Sediment Transport and Soil Detachment on Steep Slopes: I. Transport Capacity Estimation

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Precise estimation of sediment transport capacity (T_c) is critical to the development of physically based erosion models. Few data are available for estimating T_c on steep slopes. The objectives of this study were to evaluate the effects of unit flow discharge (q), slope gradient (S), and mean flow velocity on T_c in shallow flows and to investigate the relationship between T_c and shear stress, stream power, and unit stream power on steep slopes using a 5-m-long and 0.4-m-wide nonerodible flume bed. Unit flow discharge ranged from 0.625×10^{-3} to 5×10^{-3} m^2 s^{-1} and slope gradient from 8.8 to 46.6%. The diameter of the test riverbed sediment varied from 20 to 2000 μm, with a median diameter of 280 μm. The results showed that T_c increased as a power function with discharge and slope gradient with a coefficient of Nash–Sutcliffe model efficiency (NSE) of 0.95. The influences of S on T_c increased as S increased, with T_c being slightly more sensitive to q than to S. The T_c was well predicted by shear stress (NSE = 0.97) and stream power (NSE = 0.98) but less satisfactorily by unit stream power (NSE = 0.92) for the slope range of 8.8 to 46.6%. Mean flow velocity was also a good predictor of T_c (NSE = 0.95). Mean flow velocity increased as q and S increased in this study. Overall, stream power seems to be the preferred predictor for estimating T_c for steep slopes; however, the predictive relationships derived in this study need to be evaluated further in eroding beds using a range of soil materials under various slopes.

Abbreviations: ANSWERS, Areal Nonpoint Source Watershed Environment Response Simulation; NSE, coefficient of Nash–Sutcliffe model efficiency; RE, relative error; WEPP, Water Erosion Prediction Project.

The need to describe processes of soil detachment, sediment transport, and deposition for developing physically based soil erosion models has stimulated researchers to investigate new algorithms for estimating the sediment transport capacity (T_c) of shallow overland flow (Julien and Simons, 1985; Finkner et al., 1989; Govers, 1990; Ferro, 1998). Sediment transport capacity, which is the maximum equilibrium sediment load that a flow can transport, is a key concept for developing process-based erosion models because it plays a pivotal role in determining the soil detachment rate and sediment transport. A widely used modeling approach assumes that soil detachment only occurs when the sediment load (G) is less than T_c. Sediment deposition occurs when G is greater than T_c (Nearing et al., 1989). Thus, a precise estimation of sediment transport capacity is critical for successfully developing process-based soil erosion models.

At present, different hydraulic variables are used to calculate the sediment transport capacity. One of the most frequently used variables is the shear stress of shallow flow based on the bed load formula of Yalin (1963):

$$\tau = \rho g h S \qquad [1]$$

where τ is the shear stress (Pa), ρ is the water mass density (kg m^{-3}), g is the gravity constant (m s^{-2}), h is the depth of flow (m), and S is the tangent bed slope (m m^{-1}).

A few studies have been conducted in the past three decades to evaluate the performance of the Yalin equation under shallow flow conditions. Foster and Meyer (1972) evaluated several sediment transport equations with limited measured data obtained from both nonerodible flumes and plots with smaller slopes (<10%) and recommended the Yalin equation for use in simulating sediment transport in raindrop-impact overland flow. Alonso et al. (1981) thoroughly evaluated nine transport equations using 739 literature data acquired from field measurements, nonerodible flume experiments, and concave-slope tests, with sediment ranging from very fine soil particles (0.002 mm) to coarse sand (2 mm). Their results showed that the Yalin formula was the best equation for estimating the sediment transport capacity of shallow overland flow. Julien and Simons (1985) showed that the Yalin equation was less acceptable but to some extent useable for interrill sheet flow conditions. The Yalin equation has been used in the Water Erosion Prediction Project (WEPP) model for sediment transport capacity estimation (Nearing et al., 1989).

Bagnold (1966) changed the emphasis from the forces applied to the bed (shear stress) to the rate of energy expenditure, expressing the sediment transport capacity as a function of stream power per unit area of bed (Prosser and Rustomji, 2000):

$$\omega = \tau V = \rho g S q \qquad [2]$$

where ω is the stream power (kg m^{-3}), V is the mean flow velocity (m s^{-1}), and q is the unit width flow discharge (m^2 s^{-1}).

Bagnold (1980) developed a bed load equation taking into account the effects of flow depth on transport capacity using published data from both flumes and rivers in a large range of flow depths, grain size, and stream power. Li and Abrahams (1999) reported that the sediment transport capacity was positively related to excess flow power. Abrahams et al. (2001) developed a total load transport equation for interrill flow based on stream power using a large data set obtained from nonerodible flume experiments. A well-sorted sand with a median diameter of 0.74 mm was used and the flume slope varied from 5 to 17.6%. The results showed that 89.8% of the variance in transport capacity could be explained by excess flow power and flow depth. Stream power has been used to estimate transport capacity for the Griffith University Erosion System Template (GUEST) model (Yu et al., 1997).

Unit stream power became another frequently used hydraulic variable after Yang (1972, 1973) used it to develop a total load equation for cohesionless natural sands:

$$P = VS \qquad [3]$$

where P is unit stream power (m s^{-1}).

Govers (1990, 1992) proposed the use of hydraulic variables that implicitly account for the effect of bed roughness. Two suitable hydraulic variables that Govers (1992) recommended were Yang's (1972) unit stream power and Govers' (1990) effective flow power. The Govers equation, which is currently being used in the European Soil Erosion Model (EUROSEM; Morgan et al., 1998) and Limberg Soil Erosion Model (LISEM; De Roo et al., 1996), is expressed as

$$T_c = q\gamma_s m \left(P - P_c \right)^n \qquad [4]$$

where T_c is the sediment transport capacity (kg m^{-1} s^{-1}), γ_s is the mass density of the test soil (kg m^{-3}), m and n are coefficients calculated as $m = [(d_{50} + 5)/0.32]^{-0.6}$ and $n = [(d_{50} + 5)/300]^{0.25}$, P_c is the critical unit stream power (cm s^{-1}), and d_{50} is the median particle diameter of the test soil (μm).

Shear stress, stream power, and unit stream power are functions of basic hydraulic variables (such as flow rate, flow depth, mean flow velocity, and slope gradient) and cannot be measured directly. Based on dimensional analysis, Julien and Simons (1985) derived a general equation of sediment transport capacity as a power function of the flow rate, slope gradient, and rainfall intensity. The influence of rainfall intensity can be neglected when the flow depth is more than three times the raindrop diameter (Kinnell, 1988; Prosser and Rustomji, 2000). Based on a literature review of T_c and soil data, Beasley and Huggins (1982) presented a pair of equations for sediment transport capacity:

$$T_c = 146 S q^{0.5} \qquad \text{if } q \leq 0.046 \quad [5a]$$

$$T_c = 14,600 S q^2 \qquad \text{if } q > 0.046 \quad [5b]$$

where T_c is the sediment transport capacity (kg m^{-1} min^{-1}) and q is the unit width flow discharge (m^2 min^{-1}). Equations [5a] and [5b] were used in the Areal Nonpoint Source Watershed Environment Response Simulation (ANSWERS) model (Beasley and Huggins, 1982), which has been widely used during the past two decades (Ahmadi et al., 2006; Singh et al., 2006).

Slope gradient is one of the most significant factors affecting soil erosion. The characteristics of shallow flow hydraulics

and soil erosion on steep slopes are different from those on low slopes (Govers, 1992; Nearing et al., 1997, 1999; Zhang et al., 2002, 2003). Understanding sediment transport processes on steep slopes is important for the calibration and validation of process-based erosion models as well as for soil and water conservation in regions with steep terrain. In most cases, very little data exist on steep slopes. Almost all sediment transport capacity equations currently used in erosion models were developed for low slopes. For instance, the Govers equation was developed for slopes ranging from 1.8 to 14.4% (Govers, 1992). The maximum slope gradient in the experiment of Abrahams et al. (2001) was 17.6%. Specifically, little is known about erosion processes on slopes steeper than 20%, which has often been the upper limit for research in many developed countries (Liu et al., 1994; Zhang et al., 2002). Thus, it is necessary to conduct experiments on steep slopes to better understand soil erosion processes and to develop soil erosion models for use on steep slopes.

The objectives of this study were to evaluate the influence of flow discharge, slope gradient, and mean flow velocity on the sediment transport capacity under shallow flow and to examine the relationship between sediment transport capacity and shear stress, stream power, and unit stream power on steep slopes.

MATERIALS AND METHODS

The experiments were conducted at the Fangshan station of Beijing Normal University. Test materials were collected from the Yongding river bed nearby Beijing. The diameter of the test materials varied from 20 to 2000 μm (8.15% 20–100 μm, 22.09% 100–200 μm, 37.8% 200–360 μm, 23.77% 360–550 μm, and 8.19% 550–2000 μm), with a median diameter (d_{50}) of 280 μm. The average mass density of the test sediment was 2400 kg m^{-3}. The test sediment was air dried and sieved through a 2-mm sieve.

Sediment transport capacity was measured in a 5-m-long, 0.4-m-wide hydraulic flume. The bed slope of the flume could be adjusted to within 0.05% of a desired slope. The elevation of the upper flume end was adjusted by a stepping motor, allowing adjustment of the bed gradient up to 60%. The test sediment was evenly and smoothly glued on the surface of the flume bed so that the roughness (largely the grain roughness) remained constant during all the experiments.

Flow discharge was controlled by a series of valves installed on a flow diversion box and measured directly by a calibrated flow meter. Before feeding sediment, the flume bed slope and flow discharge were adjusted to the designed values. After flow became stable, measurements of flow depth were made using a level probe with an accuracy of 0.3 mm across the flow section at 0.6 m above the lower end of the flume. For each combination of flow discharge and slope gradient, 12 depths were measured. The maximum and minimum flow depths were eliminated from the data set. The average of the remaining 10 depths was considered to be the mean flow depth for that combination of flow rate and slope gradient (Table 1). Flow velocity was measured using a fluorescent dye technique in which the velocity of the leading edge of the dye was multiplied by a reduction factor of 0.8 to obtain the mean velocity (Luk and Merz, 1992). The mean values were used to calculate the shear stress, τ (Eq. [1]), the stream power, ω (Eq. [2]), and the unit stream power, P (Eq. [3]).

Two sediment sources were designed to ensure that the sediment transport capacity was reached for each combination of flow rate and slope gradient. One 1-m^3 hopper was installed over the flume at a distance of 0.5 m from the top. The sediment feeding rate was controlled by the rotating speed of rotors installed within the hopper and calibrated to measured data. The rotating speed of the rotors could be adjusted by a portable dial (con-

troller), which was adjusted manually according to the current sediment transport by flowing water. The feeding rate from the hopper for each combination of flow discharge and slope gradient was adjusted at the beginning of each test and then fixed during the test. Another sediment source was a 20-cm-wide slot across the flume bed located 0.5 m above the lower end of the flume. The slot was filled with the test sediment and was covered with a very thin iron sheet glued to the sediment with petroleum jelly to prevent water drainage during measurement of the hydraulic characteristics of the flow. After the hydraulic measurements, the hopper started to feed sediment to the flow. The sediment feeding rate was adjusted gradually until the fed sediment could not all be carried and deposition occurred near the sediment feeder, at which point the transport capacity was assumed to be reached and the feeding rate was set. The iron sheet was then removed and measurements started. If the transport capacity was not reached due to insufficient sediment feeding from hopper, the deficit would be made up by sediment entrainment from the slot to reach the sediment transport capacity. One thin iron rod was used to disturb any deposition under the hopper during the experiments. Five samples were collected for each combination of flow rate and slope gradient as quickly as possible to avoid excessive erosion in the slot. The second sediment source was refilled with test sediment if scouring occurred during the test. A new test was then started with another combination of flow rate and slope gradient.

The collected samples were allowed to settle for 24 h. The clear supernatant was decanted from the containers. The remaining wet sediment was oven dried at 105°C for 12 h. The dry sediment weight was divided by the sampling time and the flume width to obtain the sediment transport capacity. The sampling time was adjusted according to flow discharge (longer for lower flow rates and shorter for greater flow rates). The average of the five samples was used as the measured equilibrium sediment transport capacity for that combination of flow discharge and slope gradient. A series of 64 combinations of flow discharge (0.625, 1.250, 1.875, 2.500, 3.125, 3.750, 4.375, and 5.000 × 10^{-3} m^2 s^{-1}) and flume bed slope (8.8, 17.6, 22.2, 26.8, 31.5, 36.4, 41.4, and 46.6%) were tested. The measured results were also compared with the transport capacities calculated using the equations of ANSWERS and Govers.

The following statistical parameters were used to evaluate the performance of the ANSWERS and Govers transport capacity equations:

$$RRMSE = \frac{\sqrt{(1/n)\sum_{i=1}^{n}(O_i - P_i)^2}}{O_m} \quad [6]$$

where RRMSE is the relative root mean square error, O_i are the observed values, P_i are the predicted values, and O_m is the mean of the observed values;

$$RE = \frac{(P_i - O_i)}{O_i}100 \quad [7]$$

where RE (%) is the relative error;

$$NSE = \frac{\sum_{i=1}^{n}(O_i - O_m)^2 - \sum_{i=1}^{n}(P_i - O_i)^2}{\sum_{i=1}^{n}(O_i - O_m)^2} \quad [8]$$

and

$$R^2 = \frac{\left[\sum_{i=1}^{n}(O_i - O_m)(P_i - P_m)\right]^2}{\sum_{i=1}^{n}(O_i - O_m)^2 \sum_{i=1}^{n}(P_i - P_m)^2} \quad [9]$$

where R^2 is the coefficient of determination and P_m is the mean of the predicted value.

Table 1. Measured flow depths for 64 combinations of flow rates and slope gradients from 8.8 to 46.6%.

Flow rate	Flow depth							
	8.8%	17.6%	22.2%	26.8%	31.5%	36.4%	41.4%	46.6%
10^{-3} m^2 s^{-1}				mm				
0.625	1.4	1.2	1.2	1.1	1.1	1.0	1.0	0.9
1.250	2.5	2.0	1.8	1.7	1.7	1.6	1.5	1.4
1.875	3.2	2.6	2.4	2.2	2.1	2.0	1.9	1.8
2.500	3.8	3.1	2.8	2.7	2.5	2.3	2.3	2.1
3.125	4.3	3.5	3.2	3.0	2.8	2.7	2.5	2.4
3.750	4.8	3.8	3.5	3.3	3.1	2.9	2.8	2.7
4.375	5.3	4.1	3.8	3.5	3.3	3.2	3.0	2.8
5.000	5.7	4.5	4.1	3.8	3.6	3.4	3.1	3.0

RESULTS AND DISCUSSION

The measured sediment transport capacity increased as the flow discharge and slope gradient increased (Fig. 1 and 2). The relationship was a power function for both slope gradient and flow rate, with $R^2 > 0.95$. Further analysis showed that, at similar levels of shear stress, the sediment transport capacity was affected more by the flow discharge than by the slope. When the flow rate was 4.375×10^{-3} m^2 s^{-1} and the slope was 8.8%, shear stress equaled 4.54 Pa and the sediment transport capacity was 1.50 kg m^{-1} s^{-1}. When the flow rate was 1.25×10^{-3} m^2 s^{-1} and the slope was 26.8%, the flow shear stress also equaled 4.54 Pa while the measured sediment transport capacity was only 0.97 kg m^{-1} s^{-1}. The effect of slope on the sediment transport capacity increased as the slope gradient increased. For example, when the flow rate was 3.750×10^{-3} m^2 s^{-1} and the slope was 46.6%, shear stress equaled 12.14 Pa and the sediment transport capacity was 6.84 kg m^{-1} s^{-1}. When the flow rate was 4.375×10^{-3} m^2 s^{-1} and the slope was 41.4%, the flow shear stress equaled 12.01 Pa while the measured sediment transport capacity was 7.14 kg m^{-1} s^{-1}. The measured transport capacity fluctuated clearly when the slope gradient ranged from 26.8 to 36.4%, especially when flow discharge was greater (Fig. 2). The reason is unknown and further studies are needed to investigate it.
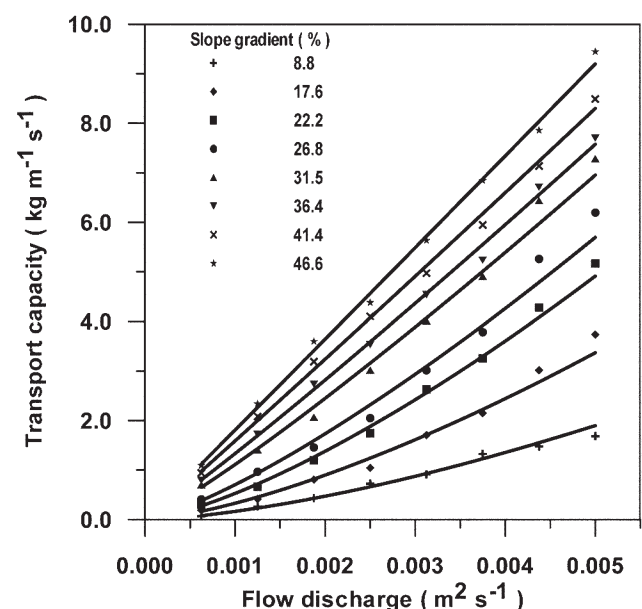

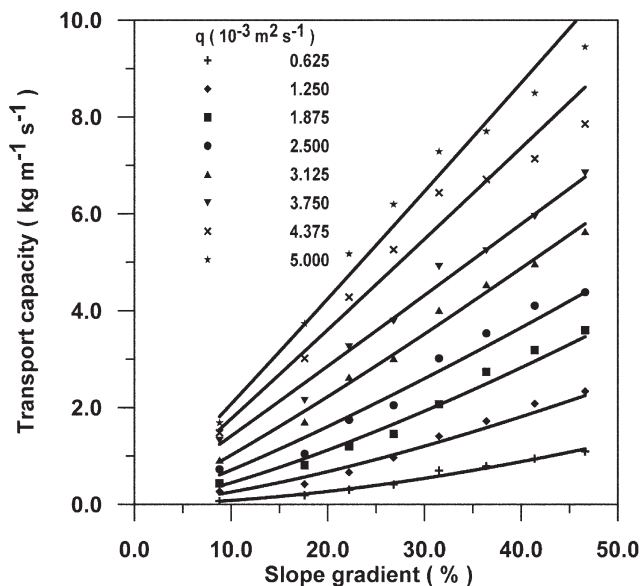**Fig. 1. Measured sediment transport capacity as a function of flow discharge at a range of slopes.**

Fig. 2. Measured sediment transport capacity as a function of slope gradient at flow rates from 0.625 to 5.000 × 10⁻³ m² s⁻¹.

Multivariate, nonlinear regression analyses between sediment transport capacity, flow discharge, and slope gradient produced the relationship of

$$T_c = 19,831 q^{1.237} S^{1.227} \quad R^2 = 0.98 \quad NSE = 0.96 \; [10]$$

where T_c is the transport capacity (kg m⁻¹ s⁻¹), q is the unit flow discharge (m² s⁻¹), and S is the slope gradient (m m⁻¹).

In general, Eq. [10] fitted the measured sediment transport capacity satisfactorily, with $R^2 = 0.98$ and NSE = 0.96. The calculated sediment transport capacities were slightly greater than the observed values, however, when the sediment transport capacity was >8.0 kg m⁻¹ s⁻¹ (Fig. 3). The exponents of unit flow discharge (1.237) and slope gradient (1.227) were within the general range of 0.9 to 1.8 as reported in the literature, but they both were less than the average value of 1.4 reported by Prosser and Rustomji (2000). The coefficient (19,831) was 35.8% greater than the coefficient of Eq. [5b] (14,600) as reported by Beasley and Huggins
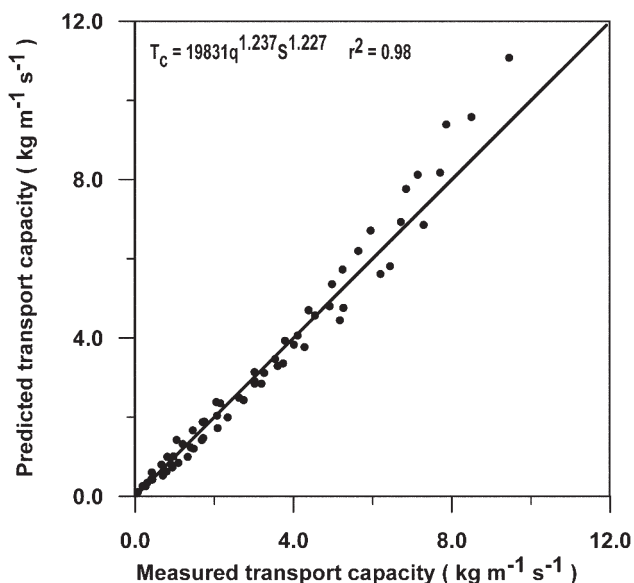
(1982), with the slope exponent being 23% greater while the discharge exponent was 39% less (1982). The differences in the exponents and the coefficient between this study and those in the literature might have largely resulted from differences in the slope gradient and sediment size distributions used in the derivation of those equations. In addition, the noneroding bed used in this study might have some effect on the derived exponents because the form roughness that influences flow depth and velocity was not simulated in this study. The exponent of q was slightly greater than that of S, however, indicating that the effect of q on T_c was slightly greater than that of S in this study, as speculated above.

Comparisons between measured and calculated sediment transport capacities using Eq. [5], as in the ANSWERS model (Beasley and Huggins, 1982), are made in Table 2 and Fig. 4. In general, the best-fit model of Eq. [10] reproduced the measured transport capacity well (Fig. 3), while Eq. [5] underestimated the measured transport capacity by about 20% (the ratio of the mean simulated transport capacity to the mean measured transport capacity [M_{Tcs}/M_{Tcm}] = 0.80). The RE predicted with Eq. [5] varied from −80 to 14%. Only five estimated values were greater than the observed counterparts. The RE was correlated with flow discharge (Fig. 5) and decreased with an increase in flow discharge. This result indicated that the simulated sediment transport capacity determined by the ANSWERS model was overly sensitive to flow discharge for the experiment data of this study.
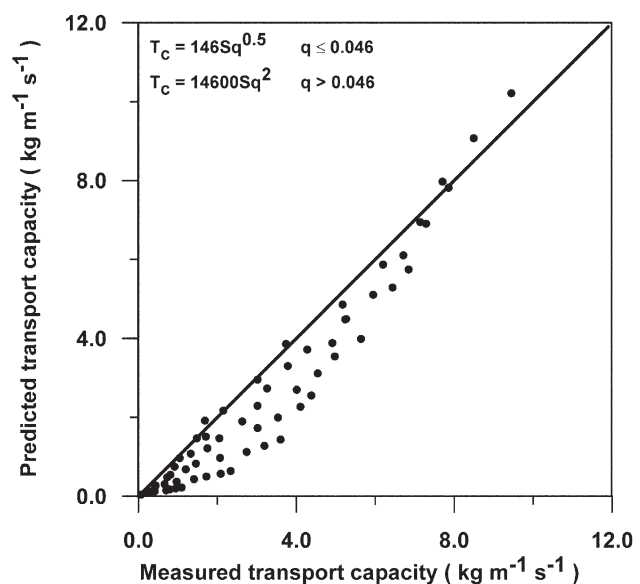
Table 2. The coefficient of determination (R^2), the coefficient of Nash–Sutcliffe model efficiency (NSE), the ratio of the mean simulated transport capacity to the mean measured transport capacity (M_{Tcs}/M_{Tcm}), the ratio of the standard deviation of simulated transport capacity to the measured transport capacity (SD_s/SD_m), the relative root mean square error (RRMSE), and the relative error (RE) statistics of simulated transport capacity using the ANSWERS and Govers models.

Model	R^2	NSE	M_{Tcs}/M_{Tcm}	SD_s/SD_m	RRMSE	RE
						%
ANSWERS	0.94	0.86	0.80	1.03	0.29	−80~14
Govers	0.96	0.70	1.17	1.44	0.24	−35~59

Fig. 3. Measured vs. predicted (using Eq. [10]) sediment transport capacity.

Fig. 4. Measured vs. predicted (using the ANSWERS model, Eq. [5a] and [5b]) sediment transport capacity.

The exponent of unit flow discharge in the ANSWERS model was 2, while the best-fit value in this study was only 1.237.

The mean flow velocity is another important factor affecting sediment transport capacity, because it is affected by both flow hydraulics (flow discharge, slope gradient, and flow depth) and surface conditions (vegetation cover and roughness). The best-fitting equation between the mean velocity and the measured sediment transport capacity was a linear function (Fig. 6):

$$T_c = 7.834V - 4.554 \quad r^2 = 0.96 \quad NSE = 0.95 \; [11]$$

where V is the mean flow velocity (m s^{-1}). From Eq. [11], the threshold velocity could be intuitively estimated when T_c was set to 0. For this study, the threshold velocity was 0.58 m s^{-1}. This implies that sediment with a median diameter of 280 μm in this study could only be transported when the mean velocity was >0.58 m s^{-1}. Many studies conducted on eroding rills have indicated that the mean flow velocity did not increase much with flow discharge and slope gradient because of the increase in form roughness as flow discharge and slope gradient increased in eroding rills (Govers, 1990; Nearing et al., 1999). This study was conducted under noneroding conditions and the mean flow velocity increased as flow discharge and slope gradient increased (Fig. 7). These differences in velocity probably have significant effects on the sediment transport capacity.

The measured sediment transport capacity was well simulated by shear stress with a power function (Fig. 8):

$$T_c = 0.054\tau^{1.982} \quad r^2 = 0.98 \quad NSE = 0.97 \; [12]$$

Compared with the WEPP's sediment transport capacity equation of $T_c = K_t\tau^{3/2}$, where K_t is a transport coefficient (Nearing et al., 1989), the exponential value of 1.982 in Eq. [12] was about 32% greater than the WEPP's exponent of 1.5. The difference in the exponents might have resulted from the facts that this experiment was conducted on steep slopes and on the noneroding bed, in which the mean flow velocity increased as q and S
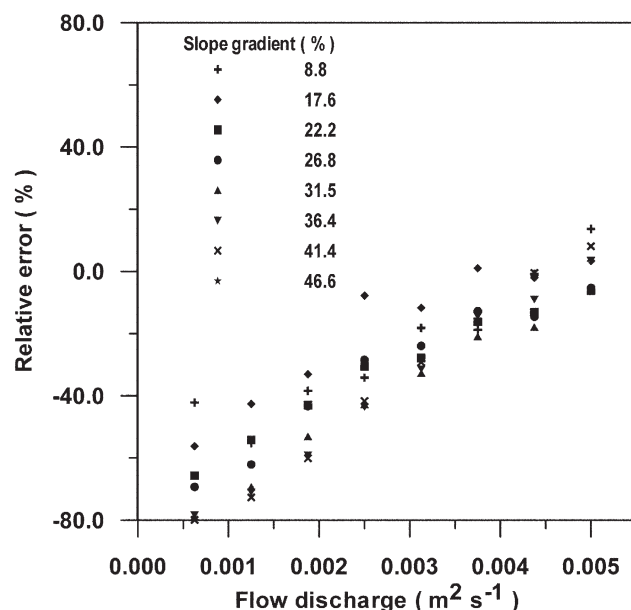
increased (Fig. 7). Govers (1992) showed that velocity increase on a rough surface was much less than the increase on a smooth surface, or it might not be influenced by bed slope on an eroding bed due to increases in the bed form roughness. A slower velocity on a rougher surface would lead to a deeper flow depth and thus a greater shear stress estimate. We speculate that a smaller exponent is needed in WEPP to offset a greater shear stress estimate for better T_c prediction. This speculation needs to be further examined and tested under eroding bed conditions for steep slopes.

The measured transport capacity was also well predicted by a linear function of stream power (Fig. 9):

$$T_c = 0.437(\omega - 0.698) \quad r^2 = 0.98 \; NSE = 0.98 \; [13]$$

Fig. 6. Measured transport capacity (T_c) as a function of mean flow velocity (V).

Fig. 5. Simulated relative error (using the ANSWERS model) vs. flow discharge across a range of slopes.

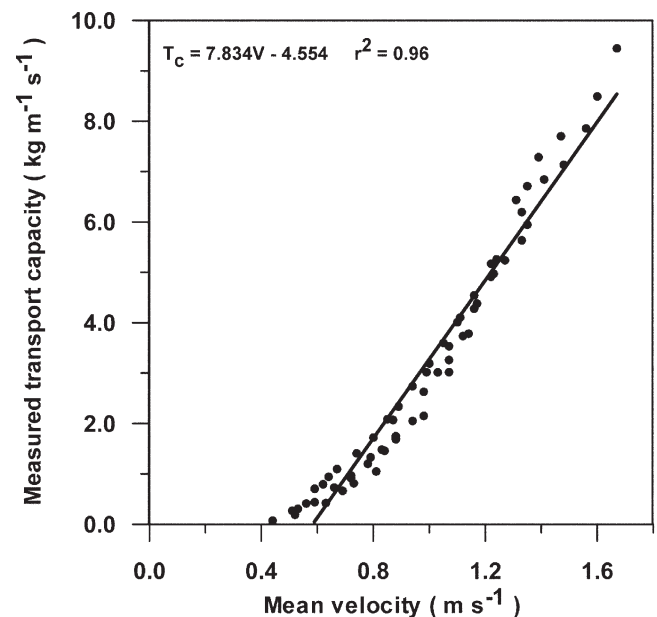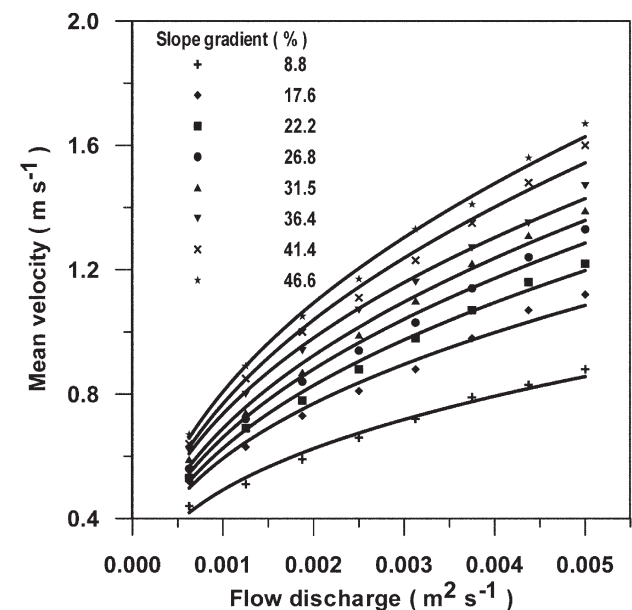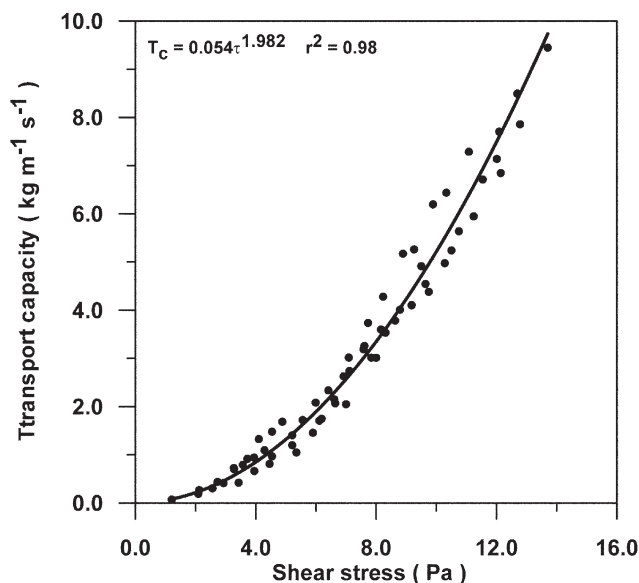**Fig. 7. Mean flow velocity as a function of flow discharge across a range of slopes.**

Fig. 8. Measured transport capacity (T_c) as a function of shear stress (τ).

Equation [13] simulated the measured transport capacity satisfactorily, with an average RE of −1.73%. The critical stream power was 0.698 for this study, which indicates that sediment could only be transported when stream power was >0.698. This result corroborates that stream power or excess stream power is a valuable hydraulic parameter for transport capacity prediction (Govers, 1990, 1992; Li and Abrahams, 1999; Abrahams et al., 2001).

The correlation between sediment transport capacity and unit stream power was not as close as either shear stress or stream power (Fig. 10):

$$T_c = 0.024 q \gamma_s \left(P - 0.4\right)^{0.863} \quad r^2 = 0.92 \ \mathrm{NSE} = 0.92 \ [14]$$

where P is the unit stream power (cm s^{-1}). The RE varied from −37 to 52%. Compared with Eq. [4], the coefficient 0.024 of Eq. [14] was greater than 0.017 by 41.2% and the exponent 0.863 was less than 0.987 by 12.6%. The fitted 0.4 equaled the data presented by Govers (1990). Equation [14] overestimated the measured trans-
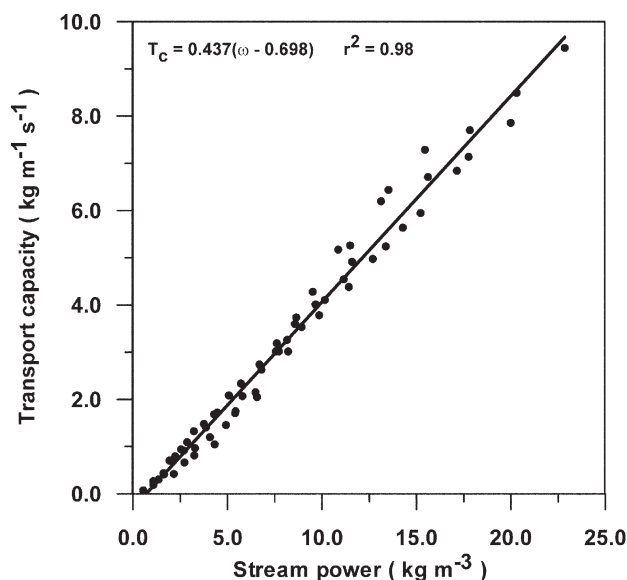

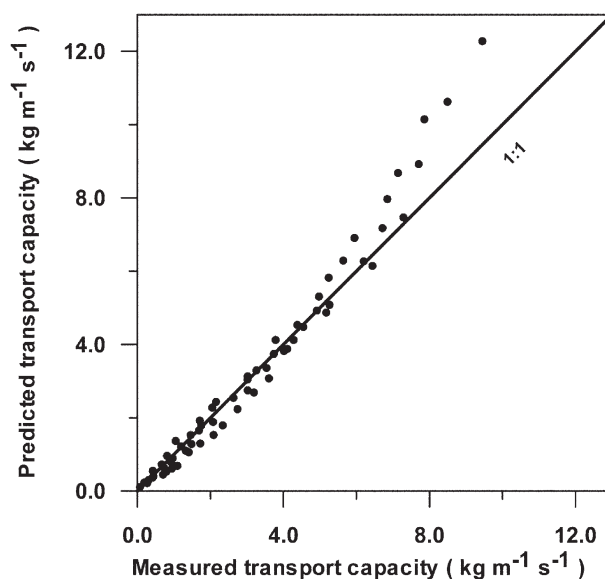**Fig. 9. Measured transport capacity (T_c) as a function of stream power (ω).**

port capacity when the transport capacity was >4.0 kg m^{-1} s^{-1} (Fig. 10). The unit stream power alone was not a good predictor for estimating T_c (note that the NSE of the best fit of Eq. [14] without q was ~0.75). Govers (1990) proposed Eq. [4], which includes a unit discharge, for T_c prediction. The simulated sediment transport capacity with the Govers equation was compared with the measured data in Fig. 11 and Table 2. The Govers equation overestimated the sediment transport capacity by about 17%. In general, Govers' equation reproduced the measured transport capacity better than the ANSWERS model (Table 2), but the error increased when the transport capacity was >4.0 kg m^{-1} s^{-1}. The RE changed from −35 to 59%. A detailed analysis indicated that the RE was correlated with both slope gradient and flow discharge. On smaller slopes (<26.8%), RE was small and no discernable trend was found between RE and flow discharge, but a distinct trend in RE was found on four steep slopes (Fig. 11). The RE varied from negative to positive as flow discharge increased, and the magnitude of the RE increased as the slope gradient increased. For example, when the slope gradient was 31.5%, the RE varied from −35 to 16%; when the slope gradient was 46.6%, the RE changed from −32 to 59% (Fig. 12). The calculated results showed that the average RE for the four smaller slopes was 13.2%, but it was 22.9% for the four steeper slopes. The difference in the experimental slope gradient probably was the principal reason for the worse prediction by the Govers equation on steep slopes.

CONCLUSIONS

The results of this study showed that sediment transport capacity increased as a power function with discharge and slope gradient; however, it was more sensitive to flow discharge than to slope gradient. The sediment transport capacity increased as a linear function of the mean flow velocity. Overall, stream power is preferred for predicting T_c because (i) it fit the measured T_c slightly better, (ii) the relationship is linear, and (iii) unit discharge and slope gradient, which are needed to compute stream power, are easy to measure or estimate.

The regressed exponential value for shear stress was 1.982, which was much greater than the WEPP's exponent of 1.5. The difference in the exponents could be caused by differences in the experimental
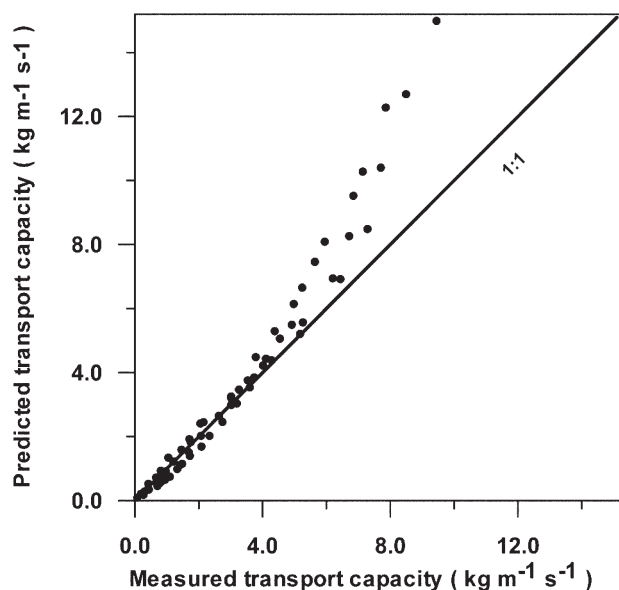
Fig. 11. Measured vs. predicted (using the Govers equation, Eq. [4]) transport capacity.

Fig. 12. Simulated relative error (using the Govers model, Eq. [4]) vs. flow discharge on steep slopes.

conditions. To further examine the difference, more studies with soil materials are needed on eroding rills and steep slopes. In general, for the two widely used transport equations (ANSWERS and Govers), the predicted sediment transport capacities matched the measured results poorly since the predicted results were overly sensitive to either flow discharge or slope gradient on steep slopes.

Based on the current study, both shear stress and stream power are good predictors for estimating T_c on steep slopes. Since this experiment was conducted using riverbed sediment (d_{50} = 280 μm) on a nonerodible flow bed, the derived relationships need to be further evaluated using soil materials on an eroding bed with steep slopes.

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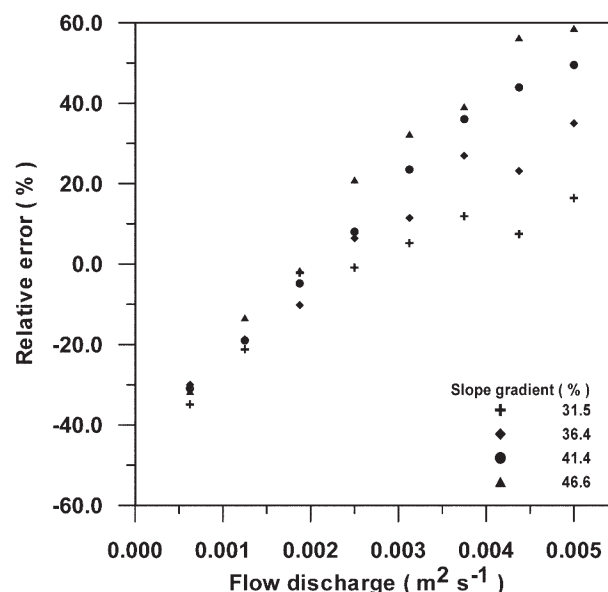Foster, G.R., and L.D. Meyer. 1972. Transport of particles by shallow flow. Trans. ASAE 19:99–102.

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